Interpretation of long-offset transient electromagnetic data from Mount Merapi, Indonesia, using a three-dimensional optimization approach

Michael Commer
Earth Sciences Division, Lawrence Berkeley National Laboratory,
Berkeley, California, USA
Formerly at Institute of Geophysics and Meteorology, University of Cologne,
Cologne, Germany

Stefan L. Helwig
Institute of Geophysics and Meteorology, University of Cologne,
Cologne, Germany

Andreas Hördt
Institute of Geology, University of Bonn,
Bonn, Germany

Bülent Tezkan
Institute of Geophysics and Meteorology, University of Cologne,
Cologne, Germany

Abstract
In the years 1998, 2000, and 2001, long-offset transient electromagnetic (LOTEM) surveys were carried out at the active volcano Merapi in Central Java. The measurements investigated the conductivity structure of the volcanic edifice. Our area of interest, which is below the summit and the upper flanks, was investigated using horizontal and vertical magnetic field time derivative data from seven transmitter-receiver setups. Because of topography and a three-dimensional (3-D) underground structure, a 3-D interpretation is used. The method optimizes few parameters of a 3-D model by a stable least squares joint inversion of the data, providing sufficient resolution capability. Reasonable data fits are achieved with a nonhorizontally layered model featuring a very conductive basement below depths of 1.5 km. While hydrothermal alteration is also considered, we tentatively explain the high conductivities by aqueous solutions with relatively high salt contents. A large magma body or a small superficial reservoir below Merapi's central volcanic complex, as discussed by other authors, cannot be resolved by the LOTEM data.1. Introduction

Introduction
Mount Merapi is a 2968 m high stratovolcano resulting from the south-north subduction of the Indian oceanic plate beneath the Eurasian plate. Merapi's topography is characteristically bell-shaped. Its base consists of a sequence of basaltic andesite lavas and intercalated pyroclastic deposits of an eroded older volcanic edifice [Camus et al., 2000]. This so-called Old Merapi is overlain by andesitic deposits of the modern Merapi. Formation and stratigraphy of the modern cone has been characterized by a succession of pyroclastic deposits and several violent and partial collapses. Presently, the activity

consists of relatively moderate pyroclastic flows and lahars accompanying growth and collapse of the active summit lava dome [Camus et al., 2000].

The LOTEM project at Merapi is a part of a multidisciplinary cooperation of the German Science Foundation (DFG) and the Volcanological Survey of Indonesia (VSI). LOTEM uses a controlled source, in this case a long horizontal electric dipole (HED), and is designed for exploration depths down to several kilometers. On the receiver side, the transient decay of the time-domain electromagnetic field, induced by transmitting a square wave current into the ground, is recorded. In this article, we interpret horizontal and vertical components of the magnetic field time derivative, referred to as fields for brevity. Horizontal fields are recorded with compact ferrite core induction coils (Zonge TEM-3). In flat terrain, we used air coils of 40 m × 40 m and 110 turns to record the vertical field. More details about the instrumentation are given by Müller et al. [2002].

The one-dimensional (1-D) interpretation of the data measured during the first LOTEM survey in 1998 revealed a resistivity distribution which is characterized by a gradual decrease with depth observed for the upper 1–1.5 km below the surface [Müller et al., 2002]. This is in good agreement with 2-D inversion results of DC resistivity measurements [Friedel et al., 2000]. Both disciplines showed a conductor with an average of 10 Ω m starting at ~1 km depth, with its isoresistivity lines generally following the topography. The 3-D magnetotelluric modeling results of Müller and Haak [2004] show a large areal extent of the conductor below the whole volcanic edifice and indicate a further resistivity decrease to values of 1 Ω m and less at depths below ~2 km. The high conductivities at such depths can possibly be related to the large-scale conductivity anomaly below Merapi identified by a regional MT survey across Central Java [Ritter et al., 1998]. This is also supported by LOTEM measurements on the lower southern flank, because at distances of up to 12 km from the volcanic cone the conductor can also be identified. These observations further indicated the transition to a different layered structure beyond ~7.5 km south of the summit, which can be approximated by a fault-like structure [Kalscheuer et al., 2004]. In the same area, Müller et al. [2002] assumed a shallow west-east striking resistivity anomaly and related it to strongly fractured remnants of an ancient avalanche caldera [Camus et al., 2000].

In this article, we focus on the gross structure of the central volcanic complex with particular interest in the region below 1.5 km depth, which could not be resolved sufficiently by the data after the 1998 survey. We have been able to improve the model significance by incorporating new data measured during 2000 and 2001 and through improved methodology. While the interpretation of the earlier results was based on 1-D inversions and qualitative data fits of single stations by 3-D forward modeling [Müller et al., 2002], we now can present the joint inversion result of data from multiple LOTEM receivers for a 3-D model of the gross structure. Using the a priori information to find a suitable model parameterization, we achieve enhanced data fits with a layer-type model incorporating the topography.

Figure 1 shows an overview of the HED transmitters and receiver stations of the data covering our area of interest. The northern HED-1 with a dipole length of 1 km was used for Stations 1–6 and is located at approximately 4 km distance from the summit. The new data (Stations 1, 1b and 2) were measured with improved equipment, enabling the recording of longer transients than possible during the 1998 survey. Station 1b, measured at the same position as Station 1, was generated by the 2 km long HED-2 at 12.8 km south from the summit. Such a two-transmitter setup significantly improves the resolution of the enclosed area. Vertical field components were measured at all receivers. In addition, horizontal fields were recorded at Stations 4, 5 and 6, thus the data set consists of 10 transients.

Methodology

The difficult terrain of the survey area makes it impossible to cover the entire volcano with profiles. The resulting sparsity of data prohibits a full large-scale 3-D inversion due to nonuniqueness. To overcome this problem, we develop a low-parameterization 3-D model of Merapi and apply a Marquardt inversion scheme, belonging to the class of unconstrained least squares methods [Marquardt, 1963]. The Marquardt scheme is typically applied to 1-D inversions of EM data, where the parameters are usually given by the resistivities and thicknesses of a horizontally layered model. Similarly, we optimize a layered 3-D model of Merapi. The finite difference (FD) modeling code of Druskin and Knizhnerman [1994] is used to calculate the 3-D model responses in the inversion. The code allows for rectangular blocks of constant resistivity, which do not have to conform to the underlying FD grid cells by means of a material averaging scheme [Moskow et al., 1999]. Illustrated in Figure 2a, this capability is used to approximate the topography with vertical columns. The entire columns are sectioned vertically such that the layer boundaries conform to the topography, as indicated by the a priori information. The inversion controls both thickness and resistivity of each layer as illustrated by the north-south oriented section in Figure 2b. The air space of the model is approximated by a high resistivity, therefore FD grid stability checks are required in order to provide reasonable model responses [Hördt and Müller, 2000]. More details about the 3-D aspects of this method can be found in the work of Commer [2003].

Preliminary studies have shown that such draped layers are superior to horizontal layers in order to achieve a satisfying data fit. The model further contains a west-east oriented division, such that the southern layer parameters vary independently from the ones in the northern section. This feature, which allows for the transition to a different type of layering in the south, might be a simplification of a more gradual north-south change. On the basis of 1-D inversion results [Kalscheuer et al., 2004], we use three layers of variable resistivity and thickness in the south, whereas four layers in the north can be found to be optimal in terms of resolution and approximation of a rather monotonic resistivity decrease with depth. Thus we invert for 12 parameters.

Results and Interpretation

Figure 2b shows the inversion result with the final layer parameters. The inversion needed 12 iterations until the decrease of the relative data misfit dropped below a predefined value. The measurements and the responses calculated from the final model are shown in Figure 3. Both the station number according to Figure 1 and the field component is given for each separate transient, where Hx and Hy denote the signals originating from the magnetic field parallel and perpendicular to the HED axis, respectively, and Hz is the vertical component. The goodness of fit χ of the actual data values di to the model predictions pi is assessed with the usual weighted least squares criterion,

where σi is the standard deviation of the ith datum. In spite of the small number of model parameters, a good data fit is achieved in general. The importance of modeling the topography is affirmed by the reproduction of the sign reversals observed in the vertical fields at Stations 5 and 6. According to the studies of Hördt and Müller [2000], such early-time reversals are caused by the 3-D nature of the mountain acting as a conductive body in the resistive air space between transmitter HED-1 and the southern flank, whereas the delay times after 10 ms are primarily influenced by deeper structures.

The Marquardt inversion is not always guaranteed to give reliable results. A solution is dependent on the data errors and the initial guess. A measure of the reliability of the result can be obtained by looking at the error bounds and the so-called importances for the estimated parameters, which can both be ascertained from a singular value decomposition analysis [Jupp and Vozoff, 1975]. The importances range from 0 to 1, which signify vanishing and maximum resolution, respectively. Table 1 shows both errors and importances for each parameter calculated from the final model result. It can be seen that the resistivities of the northern section change from well to moderately resolved toward depth, reflecting the fact that the influence of the deep layers on the model response changes toward later times, where the data noise becomes more significant. Therefore, in order to verify that the basement resistivity is still significant, we investigated the increase of χ for the data after 0.1 s due to a perturbation of the basement resistivity. For example, increasing this parameter to 0.8 Ω m (by 14%), a 20% increase of χ results. Further, the conductor must at least be 500 m thick in order to minimize χ. Below this thickness the resistivity is not resolved.

The results raise the question about the source of the high conductivities over such an areal extent as outlined in Figure 2b. It is known that melts play an important role for increased conductivities. However, a volume of the conductor's size filled with magma has not been observed at Merapi. Camus et al. [2000] related the quasi-steady magma output to a small superficial magma chamber; Ratdomopurbo and Poupinet [1995] located such a reservoir at 1.5 km below the summit. We thus have studied the effect of a simplified superficial chamber with fixed rectangular geometry as illustrated by the dashed rectangle in Figure 2b. Its conductivity is assumed to be 0.1 Ω m. The magma chamber capacity of ~1.6 × 107 m3 estimated by Gauthier and Condomines [1999] corresponds to a cube side length of ~250 m. We found that such a reservoir size causes no significant effect on the model data. To investigate whether a larger conductive reservoir is equivalent to the bottom layer in terms of the data fit, we also assumed larger side lengths of up to 2000 m and carried out further inversions where the reservoir resistivity was another model parameter. However, none of these attempts improved the data fit compared to the model without a reservoir.

Another explanation for the extensive high conductivity is that minerals produced by hydrothermal alteration lower the bulk resistivity of the rock. This has been observed in geothermal systems of other volcanoes, for example Newberry volcano [Fitterman et al., 1988] or Piton de la Fournaise [Lenat et al., 2000]. However, the extension of the conductor to the south [Kalscheuer et al., 2004], the occurrence of generally low density sediments under the volcanic deposits [Hoshino and Sunoto, 1978], and the observation that temporal gravity changes below the volcano edifice can be explained by fluid transports within a porous domain [Jentzsch et al., 2004] all suggest that the conductor contains hot saline fluids. Using estimations of porosity, salt concentration and temperature, we can find a reasonable relationship between the fluid resistivity ρw and the high bulk resistivity $\rho = 0.7\ \Omega$ m. Employing Archie's law, Le Pennec et al. [2001] studied a sample collection of clay-free recent compact angular blocks and scoria fragments from block-and-ash flow deposits of Merapi and found that the law in the form $F = \rho/\rho w = 2.35 \times \Phi{-1.2}$ provides for a satisfying relationship between the electrical formation factor F and the fractional connected porosity Φ. It is known from seismic studies that the upper subsurface layers of Merapi have a relatively low density between 2000 and 2400 kg/m3 [Wegler et al., 1999]. Porosities between 10 and 20% are assumed by Setiawan [2002] to explain observed gravity changes below the summit. Le Pennec et al. [2001] measured porosity values of 3–14% for compact angular blocks, representing the low porosity range of their studied samples. According to Nesbitt [1993], salinities in metamorphic fluids are in the range of 0–15 equivalent weight percent NaCl (eq. wt. % NaCl). Experimental studies with solutions of different salt concentrations showed a sharp decrease of ρw when increasing the temperature from 20° to 200°C and a small change when exceeding 200°C [Nesbitt, 1993]. We assume the average temperature at the depth of the conductor to be above 200°C, although much higher values are likely, because in the vicinity of the summit fumarole surface temperatures of over 600°C were measured by Zimmer et al. [2000]. To interpret the low bulk resistivity of 0.7 Ω m, a salt concentration of 25 eq. wt. % NaCl shall first be assumed. According to Nesbitt [1993], this implies a fluid resistivity of $\rho w \sim 0.008\ \Omega$ m, if the temperature exceeds 200°C. Using the relation of Le Pennec et al. [2001], a porosity of $\Phi = 5\%$ follows. In the same way, now assuming a lower salt concentration of 10 eq. wt. % NaCl involves a value of $\rho w \sim 0.02\ \Omega$ m and thus $\Phi = 10\%$. While such a bulk porosity seems to be true for the upper layers, it is more likely to decrease with depth due to compression of cracks and pores.

Conclusions

It can be concluded that the 3-D model result of Figure 2b explains the most striking features of the data, although discrepancies, in particular at the late times of Stations 5 and 6 (Figure 3), indicate the limitations due to the constrained model. Nevertheless, the reasonable data fits show that the gross resistivity structure is characterized by a dome-shaped layered model with a strong resistivity decrease with depth. The singular value decomposition analysis shows an averaged model parameter uncertainty of ~11% and a satisfying resolution of the northern layer parameters. An important feature of the model is the 0.7 Ω m conductor below ~1.5 km depth. Despite a decreasing resolution with depth, we found it still significant due to delay times later than 0.1 s in the joint data set. While a magma body with the conductor's volume seems unlikely, alteration minerals may be responsible for the high conductivities. This explanation is likely to play an important role in the hydrothermal system of Merapi. However, in view of other LOTEM and MT measurements, indicating the great extent of the conductor, and both gravity and density measurements, suggesting high porosities that are common for volcanic sediments, we also consider aqueous solutions in a porous domain as a cause of the conductor. Our estimates show that salinities significantly exceed 10 eq. wt. % NaCl and may amount to ~25 eq. wt. % NaCl in order to interpret the high conductivity.

Acknowledgements

he authors gratefully acknowledge funding of the research at Merapi by the German Science Foundation (DFG) through the program "LOTEM at Merapi" (project HO1506/8-1). We further thank VSI and MVO/BPPTK staff for support in Indonesia and V. Druskin and L. A. Knizhnerman for the permission to use their 3-D code.

References

Camus, G., A. Gourgaud, and P.-C. Mossand-Berthommier (2000), Merapi (Central Java, Indonesia): An outline of the structural and magmatological evolution, with a special emphasis to the major pyroclastic events, J. Volcanol. Geotherm. Res., 100, 139–163.[CrossRef]

Commer, M. (2003), Three-dimensional inversion of transient electromagnetic data: A comparative study, Ph.D. thesis, Univ. of Cologne, Cologne, Germany.

Druskin, V. L., and L. A. Knizhnerman (1994), Spectral approach to solving three-dimensional Maxwell's diffusion equations in the time and frequency domains, Radio Sci., 29, 937–953.[AGU]

Fitterman, D. V., W. D. Stanley, and R. J. Bisdorf (1988), Electrical structure of Newberry volcano, Oregon, J. Geophys. Res., 93, 10,119–10,134.[AGU]

Friedel, S., I. Brunner, F. Jacobs, and C. Rücker (2000), New results from DC resistivity imaging along the flanks of Merapi Volcano, in Decade-Volcanoes Under Investigation, vol. 4, edited by B. Buttkus, S. Greinwald, and J. Ostwald, pp. 23–29, Dtsch. Geophys. Ges., Hannover, Germany.

Gauthier, P.-J., and M. Condomines (1999), 210P-226Ra radioactive disequilibria in recent lavas and radon degassing: Inferences on the magma chamber dynamics at Stromboli and Merapi volcanoes, Earth Planet. Sci. Lett., 172, 111–126.[CrossRef]

Gerstenecker, C., G. Läufer, B. Snitil, and B. Wrobel (1998), Digital elevation models for Mount Merapi, in Decade-Volcanoes Under Investigation, vol. 3, edited by J. Zschau and M. Westerhaus, pp. 65–68, Dtsch. Geophys. Ges., Hannover, Germany.

Hördt, A., and M. Müller (2000), Understanding LOTEM data from mountainous terrain, Geophysics, 65, 1113–1123.[CrossRef]

Hoshino, K., and W. Sunoto (1978), Geology-geologic structures, in Gravity and Geologic Studies in Java, Indonesia, Spec. Publ. Ser., vol. 6, edited by M. Untung and Y. Sato, pp. 136–144, Geol. Surv. of Indonesia, Bandung.

Jentzsch, G., A. Weise, C. Rey, and C. Gerstenecker (2004), Gravity changes and internal processes: Some results obtained from observations at three volcanoes, Pure Appl. Geophys., 161, 1415–1431.[CrossRef]

Jupp, D. L. B., and K. Vozoff (1975), Stable iterative methods for the inversion of geophysical data, Geophys. J. R. Astron. Soc., 42, 957–976.

Kalscheuer, T., S. L. Helwig, B. Tezkan, and M. Commer (2004), Multidimensional modelling of LOTEM data collected at the south flank of Mt. Merapi, Indonesia, in Protokoll über das 20. Kolloquium elektromagnetische Tiefenforschung, edited by A. Hördt and J. Stoll, pp. 104–113, Dtsch. Geophys. Ges., Potsdam, Germany.

Lenat, J. F., D. V. Fitterman, D. B. Jackson, and P. Labazuy (2000), Geo-electrical structure of the central zone of Piton de la Fournaise Volcano (Réunion), Bull. Volcanol., 62, 75–89.[CrossRef]

Le Pennec, J.-L., D. Hermitte, I. Dana, P. Pezard, C. Coulon, J.-J. Cochemé, E. Mulyadi, F. Ollagnier, and C. Revest (2001), Electrical conductivity and pore-space topology of Merapi lavas: Implications for the degassing of porphyritic andesite magmas, Geophys. Res. Lett., 28, 4283–4286.[AGU]

Marquardt, D. W. (1963), An algorithm for least-squares estimation of non-linear parameters, SIAM J., 11, 431–441.[CrossRef]

Moskow, S., V. L. Druskin, T. Habashy, P. Lee, and S. Davydychewa (1999), A finite difference scheme for elliptic equations with rough coefficients using a cartesian grid nonconforming to interfaces, SIAM J. Numer. Anal., 36, 442–464.[CrossRef]

Müller, A., and V. Haak (2004), 3-D modeling of the deep electrical conductivity of Merapi volcano (Central Java): Integrating magnetotellurics, induction vectors and the effects of steep topography, J. Volcanol. Geotherm. Res., 138, 205–222.[CrossRef]

Müller, M., A. Hördt, and F. M. Neubauer (2002), Internal structure of Mount Merapi, Indonesia, derived from long-offset transient electromagnetic data, J. Geophys. Res., 107(B9), 2187, doi:10.1029/2001JB000148.[AGU]

Nesbitt, B. E. (1993), Electrical resistivities of crustal fluids, J. Geophys. Res., 98, 4301–4310.[AGU]

Ratdomopurbo, A., and G. Poupinet (1995), Monitoring a temporal change of seismic velocity in a volcano: Application to the 1992 eruption of Mt. Merapi (Indonesia), Geophys. Res. Lett., 22, 775–778.[AGU]

Ritter, O., A. Hoffmann-Rothe, A. Müller, S. Dwipa, E. M. Arsadi, A. Mahfi, I. Nurnusanto, S. Byrdina, F. Echternacht, and V. Haak (1998), A magnetotelluric profile across Central Java, Indonesia, Geophys. Res. Lett., 25, 4265–4268.[AGU]

Setiawan, A. (2002), Modeling of gravity changes on Merapi volcano observed between 1997–2000, Ph.D. thesis, Tech. Univ. of Darmstadt, Darmstadt, Germany.

Wegler, U., N. Maercklin, B. G. Lühr, W. Rabbel, and J. Zschau (1999), Scattering and absorption of seismic waves at Merapi volcano, paper presented at EUG 10, Eur. Union of Geosci., Strasbourg, France.

Zimmer, M., J. Erzinger, and Y. Sulistiyo (2000), Continuos chromatographic gas measurements on Merapi volcano, Indonesia, in Decade-Volcanoes Under Investigation, vol. 4, edited by B. Buttkus, S. Greinwald, and J. Ostwald, pp. 87–91, Dtsch. Geophys. Ges., Hannover, Germany.